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Holocene sediment records from the continental shelf of Mac. Robertson Land, East Antarctica

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Abstract. Geochemical records are presented for five sediment cores from basins on the continental shelf of Mac. Robertson Land, East Antarctica. The cores contain 2-4 m thick sequences of hemipelagic, siliceous mud and ooze (SMO) deposited under seasonally open marine conditions. The inner and middle shelf SMO sequences are massive dark olive green material, whereas the outer shelf SMO sequences are dark olive material interspersed with light olive green layers ~1-10 cm thick. The biogenic material is dominated by marine diatoms including *Fragilariopsis curta*, *Fragilariopsis cylindrus*, and *Chaetoceros* spp. in the dark-colored SMO and *Corethron criophilum* in the light-colored layers. Radiocarbon dates suggest that the cores provide continuous accumulation records extending from < 1 kyr before present (B.P.) back as far as 4-15 kyr B.P., with estimated accumulation rates of 0.07-5 mm yr^{-1}. The three core records from the middle and outer shelf suggest six episodes of increased accumulation of biogenic material at ~5.5 kyr B.P. (all three cores), 1, 2, and 6.2 kyr B.P. (two of the three cores), and 3.8 and 10.8 kyr B.P. (one core), most of which coincide with *Corethron* layers. We interpret these features as the result of enhanced diatom production over the outer shelf, possibly related to climatic warm periods. The absence of such features in the inner shelf core records is thought to reflect a relatively constant level of seasonal diatom production in adjacent waters maintained by a coastal polynya.

1. Introduction

The Antarctic continental shelf accounts for a significant fraction of Southern Ocean primary production and is a major area of oceanic deepwater formation [*Comiso et al.*, 1993; *Arrigo et al.*, 1998a; *Deacon*, 1984; *Orsi et al.*, 1999]. Algal production in Antarctic shelf waters may thus play a significant role in the biogeochemical cycles of carbon and silicon, and in defining the composition of oceanic bottom waters [*Smith and Gordon*, 1997; *Nelson et al.*, 1996; *Arrigo et al.*, 1999]. At present, little is known about algal production and its relationship to environmental conditions on the Antarctic shelf during the late Quaternary. This largely reflects the dynamic nature of this continental margin, where seafloor sediments are widely reworked and redistributed by the action of ice and currents [*Dunbar et al.*, 1985; *Anderson and Molnia*, 1989; *Harris and O'Brien*, 1996; *Anderson*, 1999]. However, some fjords and shelf basins provide natural sediment traps, where there are accumulations of hemipelagic sediments derived from overlying waters and adjacent shelf areas [*Domack*, 1982; *Domack and McClennen*, 1996; *Harris and O'Brien*, 1996; *Barker et al.*, 1998; *Harris and O'Brien*, 1998]. A number of studies have made use of sediment cores from such locations to infer paleoenvironmental conditions on the Antarctic shelf during the Holocene and late Pleistocene [e.g., *Leventer and Dunbar*, 1988; *Domack et al.*, 1993; *Leventer et al.*, 1993, 1996; *Shevenell et al.*, 1996; *Frignani et al.*, 1998; *Sedwick et al.*, 1998; *Cunningham et al.*, 1999; *Domack and Mayewski*, 1999; *Domack et al.*, 1999].

In these investigations, down core chemical, physical, and micropaleontological data together with radiocarbon chronologies have been used to construct regional records of the relative accumulation of biogenic versus lithogenic material, from which paleoenvironmental conditions have been inferred. A mid-Holocene climatic warming has been postulated on the basis of sediment records from the continental shelf of George V and Adélie Land, Prydz Bay, and the western Ross Sea [*Domack et al.*, 1991; *Jacobson*, 1997; *Frignani et al.*, 1998; *Cunningham et al.*, 1999], whereas sediments from the western margin of the Antarctic Peninsula and the continental shelf of Mac. Robertson Land contain evidence of century- to millennium-scale variations in accumulation of biogenic matter during the Holocene [*Domack et al.*, 1993; *Domack and Mayewski*, 1999; *Leventer et al.*, 1996; *Sedwick et al.*, 1998]. Such changes have also been inferred from a high-resolution Ocean Drilling Program sediment record from the Palmer Deep on the Antarctic Peninsula, which contains evidence of ~400, 200, and 50-70 year cycles in accumulation of pelagic biogenic material as well as longer-term paleoenvironmental changes, including a late Holocene neoglacial period, a mid-Holocene climatic optimum, an early Holocene climatic cooling, and a late Pleistocene deglacial episode [*Domack et al.*, 2001].

These results naturally raise questions concerning the regional coherence of such records and the spatial scale of the inferred paleoenvironmental variations, given that present-day environmental conditions on the Antarctic margin, such as sea ice cover and algal biomass, are known to be highly variable [*Comiso et al.*, 1993; *Arrigo et al.*, 1998b; *Barker et al.*, 1998; *Parkinson*, 1998]. Such questions can only be addressed by examining the

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coherence of paleoenvironmental records from the Antarctic shelf over a wide range of spatial scales. Here we present geochemical and sedimentological records for five sediment cores collected from shelf basins on the continental margin of Mac. Robertson Land, East Antarctica (the Mac. Robertson Shelf), a region which sustains relatively high algal biomass [*Comiso et al.,* 1993] and may be a significant source of Antarctic Bottom Water [*Orsi et al.,* 1999]. Preliminary analyses of two of these cores indicated significant differences in the sedimentary records from the inner and outer shelf over distances of < 100 km [*Sedwick et al.,* 1998]. The new data presented here indicate that there have indeed been significant small-scale regional variations in the accumulation of biogenic material in the Mac. Robertson Shelf basins during the Holocene, but our results also provide evidence for millennial-scale environmental variations in common with sedimentary records from other parts of the Antarctic margin.

2. Materials and Methods

2.1. Study Area

The Mac. Robertson Shelf extends some 400 km west of Prydz Bay, East Antarctica, with a typical width of 90 km (Figure 1). Here the continental shelf is made up of relatively shallow banks < 200 m in depth, separated by steep-sided basins and valleys up to 1200 m in depth which are interpreted as relict glacial troughs [*O'Brien, et al.,* 1994; *Harris and O'Brien,* 1998]. Dense, high-salinity shelf water forms along the shelf in association with coastal polynyas, and this water sinks and flows off the shelf, probably filling the perched basins [*Baines and Condie,* 1998; *Harris and O'Brien,* 1998; *Harris,* 2000]. On the outer shelf and upper continental slope a large-scale westward flowing current follows the Antarctic Slope Front, extending from the surface waters to the seafloor, where current speeds up to several meters per second have been measured [*Smith et al.,* 1984; *Harris and O'Brien,* 1998; *Bindoff et al.,* 2000]. Coarse-grained sand and gravel deposits occur on the shallower areas of the outer shelf and upper slope, whereas finer-grained muds occur in the deep basins, particularly on the inner shelf [*Harris and O'Brien,* 1996, 1998]. These basins contain up to several meters (at least) of siliceous mud and ooze (SMO) deposited under seasonally open marine conditions, overlying sandy silt and glacial marine muds that were probably deposited under or near a permanent ice shelf [*Harris and O'Brien,* 1998; *Anderson,* 1999]. The radiocarbon ages of these facies in sediment cores from the Mac. Robertson shelf suggest that seasonally open marine conditions commenced around 10-12 kyr before present (B.P.) on the outer shelf and around 6 kyr B.P. on the inner shelf, the difference representing the period over which a permanent ice canopy retreated across the continental shelf during the early Holocene [*Harris and O'Brien,* 1998; *Sedwick et al.,* 1998].

2.2. Sample Materials

The sediment cores used in this study were collected from the Nielsen Basin (maximum water depth ~1200 m) and Iceberg Alley (maximum water depth ~500 m), which are two of the larger basins on the Mac. Robertson Shelf (Figure 1). The 9 cm diameter gravity cores were collected during cruises of RSV *Aurora Australis* in 1993, 1995, and 1997. Five cores were examined in this study (approximate water depths in parentheses): AA186-GC34 (470 m) and KROCK-GC1 (478 m) from the outer shelf in Iceberg Alley, AA149-GC2 (1100 m) and KROCK-GC2 (1090 m) from the inner shelf in the Nielsen Basin, and AA149-GC12 (626 m) from midshelf in the Nielsen Basin (Figure 1). These sites are separated by distances ranging from ~5 to 100 km. All of the cores contain continuous sequences of SMO ranging in thickness from 267 to 374 cm (Figure 2). In cores KROCK-GC2 and AA149-GC12 the SMO units overlie sandy glacial marine muds. The SMO sequences are primarily a mixture of diatom ooze, fine-sand to coarse-silt quartz, and other fine lithogenic material. The SMO units of the

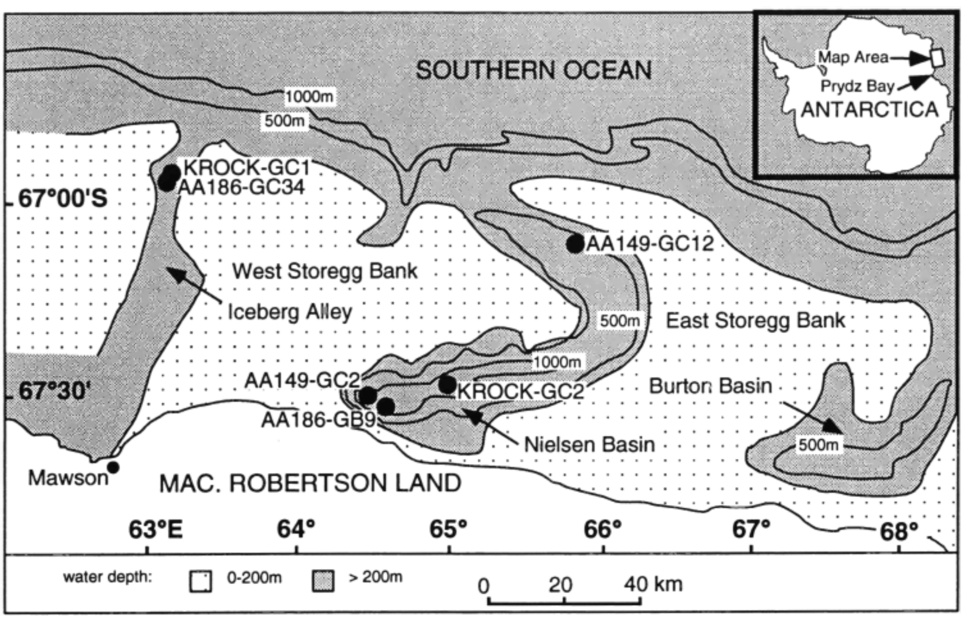


Figure 1. Map of the Mac. Robertson Shelf, showing sediment collection sites.

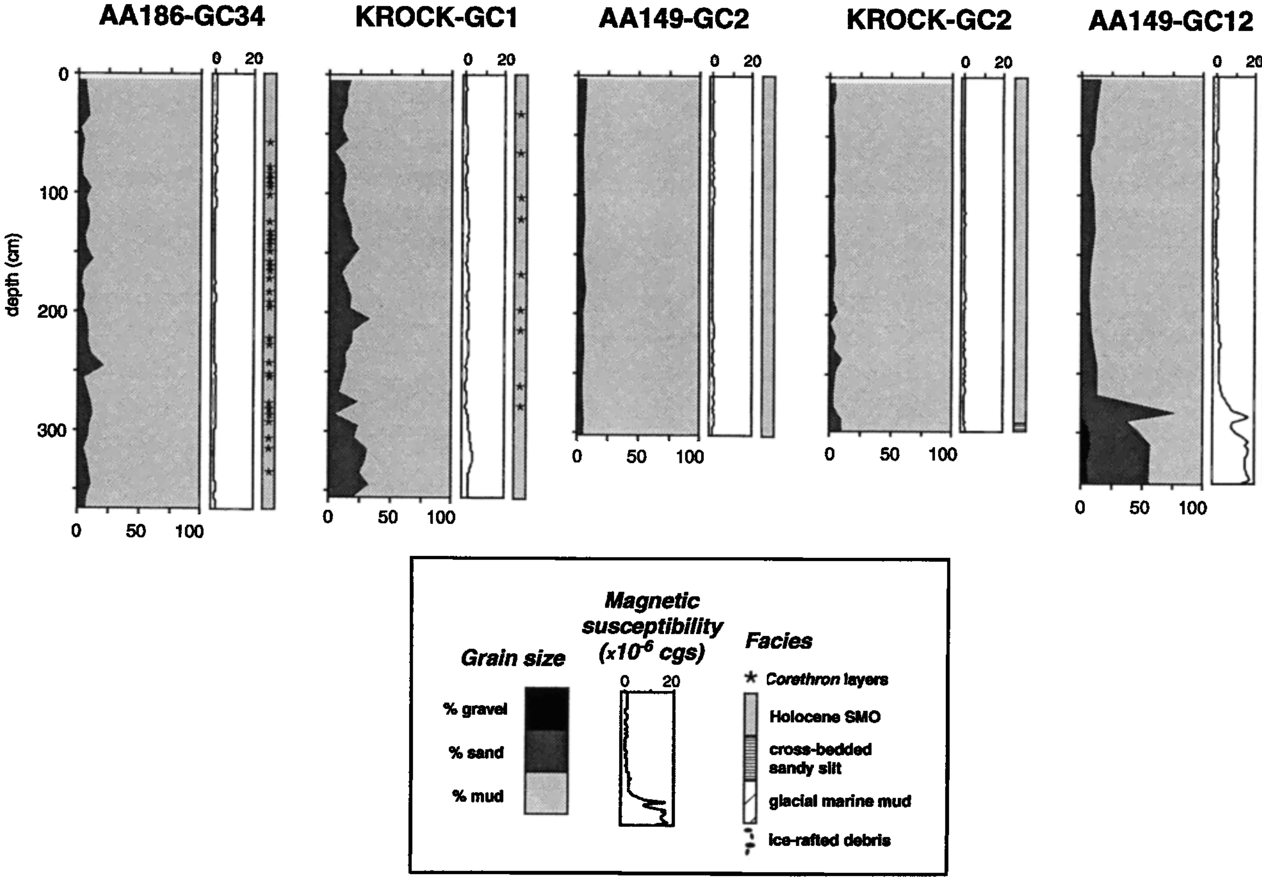


Figure 2. Grain size, magnetic susceptibility and facies classification for cores in this study.

inner and middle shelf cores (AA149-GC2, KROCK-GC2, and AA149-GC12) are massive, generally featureless, dark olive green sediment, whereas the SMO units from the outer shelf cores (AA186-GC34 and KROCK-GC1) are dark olive green material interspersed with fluffy, light olive green bands, which range in thickness from ~1 to 10 cm. Finer-scale light and dark laminations were apparent in cores AA186-GC34 and KROCK-GC1 when they were first split, but these features faded after several weeks of storage.

The cores were split and described immediately after collection, then wrapped in polyethylene and stored at 2°C. The cores were subsampled for various chemical and physical measurements in Hobart. Subsamples were taken over 10 cm intervals for geochemical analysis, and the geochemical data presented in section 3.2 represent the depth-averaged bulk compositions of these 10 cm thick subsamples. In addition, 1-2 cm thick subsamples were taken from selected depths for radiocarbon dating and grain-size analysis, 1 cm thick subsamples were taken from the upper 5-10 cm of the cores for gamma spectrometric analysis, and 1 cm^3 subsamples were taken at 10 cm intervals for determination of dry bulk density. The uppermost portions of the outer shelf cores suffered some compaction (< 10 cm) and minor stratigraphic disturbance after collection owing to the high water content of the sediment, but we have made no attempt to correct our data for these effects. In addition, the uppermost section of core KROCK-GC1 shrank in length from 100 to ~90 cm prior to subsampling owing to water loss during storage. In the presentation of data for this core, average sample depths between 0 and 90 cm have been multiplied by a factor of 100/90 in an effort to correct for this shrinkage.

2.3. Core Chronology and Isotopic Analyses

The near absence of carbonate microfossils in the cores precludes use of standard $\delta^{18}O$ stratigraphy and/or radiocarbon dating of calcium carbonate, whereas the likelihood of non-uniform sedimentation rates and the presence of significant concentrations of authigenic uranium (see section 3.2) precludes the estimation of accumulation rates using the uranium series radionuclides ^{226}Ra, ^{230}Th, and ^{231}Pa. The primary chronostratigraphic tool we have used in this study is radiocarbon dating of bulk organic carbon (typically 1-2% by mass in these sediments), which has been successfully employed in other studies of sediments from the Antarctic continental shelf [e.g., *Domack et al.*, 1989; *Leventer et al.*, 1996; *Domack et al.*, 2001]. Radiocarbon ages were determined by accelerator mass spectrometry at either the Australian Nuclear Science and Technology Organization (ANSTO) or the New Zealand Institute of Geological and Nuclear Sciences (NZI). Radiocarbon dates are reported here as conventional radiocarbon years before present, as defined by *Stuiver and Polach* [1977]. The $\delta^{13}C$ values used to calculate the radiocarbon ages were either measured (NZI analyses) or assumed (ANSTO analyses), with the assumed values based on $\delta^{13}C$ measurements of subsamples from cores KROCK-GC1 and KROCK-GC2 performed by the Australian Geological Survey Organization. The errors introduced in the radiocarbon ages due to the use of assumed $\delta^{13}C$ values are likely to be less than the analytical uncertainties in the ^{14}C measurements, given the range of measured $\delta^{13}C$ values (-24.1 to –34.1‰). In addition, unsupported ^{210}Pb was determined in core top subsamples by gamma spectrometry [*McMurtry et al.*, 1995] in an effort to evaluate the possible loss of core top material during sample collection.

2.4. Physical and Geochemical Measurements

Down core magnetic susceptibility, which provides a relative measure of ferromagnetic (i.e., lithogenic) mineral content [*Leventer et al.*, 1996], was determined with a Bartington MS-2 magnetic susceptibility meter. The cores were also X-rayed in order to identify macroscopic sedimentary structures, and down core subsamples were wet sieved to determine percentage gravel, sand, and mud by dry weight [*Harris and O'Brien*, 1998]. The following geochemical measurements were performed on the 10 cm thick subsamples: (1) bulk major and minor elements (Al, Si, Ti, Mn, Fe, Ni, Cu, Zn, Br, Mo, Ba, and U) were determined in crushed, 60°C dried (and, for major elements, deionized water washed) material by X-ray fluorescence spectroscopy, following a modification of the method of *Shimmield* [1984]; (2) biogenic silica (opal), with assumed composition $SiO_2 \cdot 0.4H_2O$, was determined in freeze-dried sediment by the method of *Mortlock and Froelich* [1989]; and (3) total organic carbon (TOC) was determined in crushed, 60°C dried, acid-treated, deionized water washed material using an elemental analyzer at the Australian Geological Survey Organization or the University of Tasmania. Analytical uncertainties, as estimated from repeated measurements of in-house standards, are presented in Table 1 (the geochemical data presented here are available electronically at http://www.antcrc.utas.edu.au/antcrc/research/sediment_web/data/geochem.html).

Table 1. Estimated Analytical Uncertainties

Species	Uncertainty[a]
Al	0.5
Si	0.2
Ti	5
Mn	10
Fe	2
Ni	10
Cu	10
Zn	1
Br	10
Mo	5
Ba	1
U	20
Opal	5
TOC	0.2[b]

[a]Relative standard deviation on mean.
[b]Absolute standard deviation (wt %).

3. Results

3.1. Core Preservation and Chronostratigraphy

Table 2 presents the radiocarbon ages and measured or assumed $\delta^{13}C$ values of bulk organic carbon in subsamples from the five cores. The measured $\delta^{13}C$ values are generally consistent with the range of –20 to –30‰ reported for Southern Ocean pelagic phytoplankton [*Gibson et al.*, 1999; *Popp et al.*, 1999], although slightly lower values (< -32‰) in the upper portion of core AA149-GC12 may reflect the presence of ^{13}C-depleted relict terrestrial organic matter [*Harris et al.*, 1996]. Down core radiocarbon ages generally increase in a regular fashion (Figure 3) and suggest that the cores preserve continuous records of sediment accumulation over time periods ranging from ~3.8 kyr (AA149-GC2) to 15 kyr (AA149-GC12). Our initial analyses of core top samples from cores KROCK-GC1 and KROCK-GC2 detected no unsupported ^{210}Pb [*Sedwick et al.*, 1998], suggesting the loss of sediments corresponding to the past ~100-200 years (at least) of accumulation during collection of these cores. Subsequent analyses (data not shown) indicate low levels of unsupported ^{210}Pb in only the upper 2 cm of cores AA186-GC34 and KROCK-GC2, and no unsupported ^{210}Pb in core KROCK-GC1, consistent with some loss of core top material during collection, whereas the low to moderate levels of unsupported ^{210}Pb measured in the upper 5 cm of cores AA149-GC2 and AA149-GC12 suggest that there were no significant core top losses.

In our study region the radiocarbon age of organic matter at the sediment-water interface is expected to be greater than zero as a result of (1) the nonzero radiocarbon age of the dissolved inorganic carbon that is converted into organic matter in the euphotic zone (assumed to be the principal source of organic carbon in our sediment cores), termed the reservoir effect, which is ~1300 years in surface waters of the Southern Ocean [*Gordon and Harkness*, 1992; *Berkman and Forman*, 1996]; (2) bioturbation in the upper sediment column, which vertically mixes material over depths of the order of 10 cm [*Berner*, 1980; *Libes*, 1992]; and (3) dilution of fresh sediments by older, resuspended particulate carbon [*Harris et al.*, 1996]. In an effort to correct our radiocarbon-based chronologies for the combined effects of these processes, we have subtracted 1730 radiocarbon years from our raw radiocarbon ages. This value of 1730 radiocarbon years is the raw radiocarbon age of a well-stratified, water-saturated, surface sediment grab sample (AA186-GB9) that was recovered near the location of AA149-GC2 in the inner Nielsen Basin (E. Domack, personal communication, 1997). The radiocarbon age of this surface sediment sample is assumed to be representative of surface sediments within the shelf basins of our study area.

We recognize that there are a number of significant uncertainties included in our radiocarbon age correction. One is the possible geographic variation in radiocarbon age of surface sediments in these shelf basins, which might be expected, for example, because of differences in the proportion of resuspended material accumulating at different locations. Another uncertainty is introduced by the likely variation in the reservoir age of Antarctic waters between the Last Glacial Maximum and the early Holocene, which may be of the order of thousands of years, based on our knowledge of changes in the radiocarbon age of oceanic deep waters over this period [*Samson*, 1999; *Sikes et al.*, 2000] and given that upwelled deep waters dominate the radiocarbon inventory of Antarctic surface waters [*Berkman and Forman*, 1996]. Yet an additional complication to the radiocarbon age correction is introduced by the presence of bomb-

Table 2. Radiocarbon Ages

Sample	Depth, cm	$\delta^{13}C$, ‰	Age,[a] years B.P.	Corrected Age,[b] years B.P.	Analysis Number[c]
AA186-GB9/0-2	1	-24.3	1733 ± 83	0	NZA 7716
AA186-GC34/0-1	0.5	-25.8	1987 ± 83	254	NZA 7712
AA186-GC34/180-181	180.5	-30.5	3881 ± 83	2148	NZA 7713
AA186-GC34/272-273	272.5	-25.7	5611 ± 84	3878	NZA 7714
AA186-GC34/367-368	367.5	-26.1	7949 ± 84	6216	NZA 7715
KROCK-GC1/0-1	0.5	(-25)[d]	2630 ± 80	897	OZB 995
KROCK-GC1/85.5-86	99.5	-24.9	3838 ± 84	2105	NZA 7717
KROCK-GC1/85.5-86.5	99.5	(-25)	3980 ± 80	2247	OZB 996
KROCK-GC1/181-182	181.5	(-25)	5940 ± 80	4207	OZB 997
KROCK-GC1/272-273	272.5	(-25)	7200 ± 130	5467	OZB 998
KROCK-GC1/356.5-357.5	357	-24.14	13390 ± 150	11657	NZA 4639
AA149-GC2/0-1	0.5	-24.5	1786 ± 71	53	NZA 5779
AA149-GC2/35-36	35.5	(-25)	1710 ± 100	0	OZC 080
AA149-GC2/70-71	70.5	(-25)	2250 ± 70	517	OZC 081
AA149-GC2/105-106	105.5	(-25)	3040 ± 140	1307	OZC 082
AA149-GC2/140-141	140.5	(-25)	2390 ± 110	657	OZC 083
AA149-GC2/175-176	175.5	(-25)	2910 ± 70	1177	OZC 084
AA149-GC2/210-211	210.5	(-25)	3350 ± 70	1617	OZC 085
AA149-GC2/245-246	245.5	(-25)	4070 ± 90	2337	OZC 086
AA149-GC2/273.5-275.5	274.5	(-25)	4420 ± 150	2687	OZC 087
AA149-GC2/302-303	302.5	-24.4	5498 ± 88	3765	NZA 5782
KROCK-GC2/7.5-8.5	0.5	(-24.4)	2030 ± 310	297	OZB 098
KROCK-GC2/42.5-43.5	35.5	(-25)	1940 ± 70	207	OZC 076
KROCK-GC2/77.5-78.5	70.5	(-24.2)	2420 ± 80	687	OZB 099
KROCK-GC2/112.5-113.5	105.5	(-25)	2750 ± 60	1017	OZC 077
KROCK-GC2/147.5-148.5	140.5	(-23.9)	3330 ± 100	1597	OZB 100
KROCK-GC2/182.5-183.5	175.5	(-25)	3950 ± 100	2217	OZC 078
KROCK-GC2/217.5-218.5	210.5	(-23.9)	5060 ± 180	3327	OZB 101
KROCK-GC2/252.5-253.5	245.5	(-25)	5970 ± 150	4237	OZC 079
KROCK-GC2/281-283	274.5	-26.87	7673 ± 84	5940	NZA 4640
AA149-GC12/2-3	2.5	-25.6	2171 ± 66	438	NZA 5964
AA149-GC12/40-41	40.5	-34.1	5519 ± 71	3786	NZA 6754
AA149-GC12/80-81	80.5	-32.3	5380 ± 78	3647	NZA 6755
AA149-GC12/120-121	120.5	-29.6	7124 ± 77	5391	NZA 6756
AA149-GC12/200-201	200.5	-28.1	8102 ± 81	6369	NZA 6749
AA149-GC12/250-251	250.5	-25.5	11410 ± 110	9677	NZA 6063
AA149-GC12/265-266	265.5	-25.1	12122 ± 92	10389	NZA 7718
AA149-GC12/301-302	300.5	-26.3	17150 ± 280	15417	NZA 5763

[a]Radiocarbon years B.P. as defined by *Stuiver and Polach* [1977].
[b]Corrected age assumes surface sediment age of 1733 radiocarbon years B.P.
[c]Sample numbers beginning with OZ analyzed by the Australian Nuclear Science and Technology Organization, and sample numbers beginning with NZ analyzed by the New Zealand Institute of Geological and Nuclear Sciences.
[d]Values in parentheses are assumed for the calculation of radiocarbon ages.

derived radiocarbon in sediments deposited during the past 50 years, which will have the effect of decreasing the radiocarbon age of recent sediments by as much as 500 years [*Berkman and Forman*, 1996]. However, in the absence of specific information regarding the effect of these various processes on the radiocarbon age of surface sediments in basins on the Mac. Robertson Shelf, we have corrected all of our raw radiocarbon ages by subtracting 1730 radiocarbon years. The thus corrected ages of our sediment core samples, which we compare against the conventional radiocarbon timescale, are presented in Table 2.

Sediment accumulation rates have been calculated between dated samples in the cores as shown in Figure 3. These calculated sedimentation rates range from 7 to 500 cm kyr^{-1} (0.07-5 mm yr^{-1}), although the highest rates, calculated for the uppermost sections of the cores, have large uncertainties resulting from the analytical uncertainties in the radiocarbon ages. Typical accumulation rates appear to be of the order of 50 cm kyr^{-1}. Several dated samples were excluded from the accumulation rate calculations in cases where either (1) the sample age is older than the age of the sample immediately below (cores AA149-GC2 and AA149-GC12), suggesting the possible input of older material due to slumping, or (2) the sample age is within analytical uncertainty of the age of the core top sample immediately above (cores AA149-GC2 and KROCK-GC2). The corrected radio-

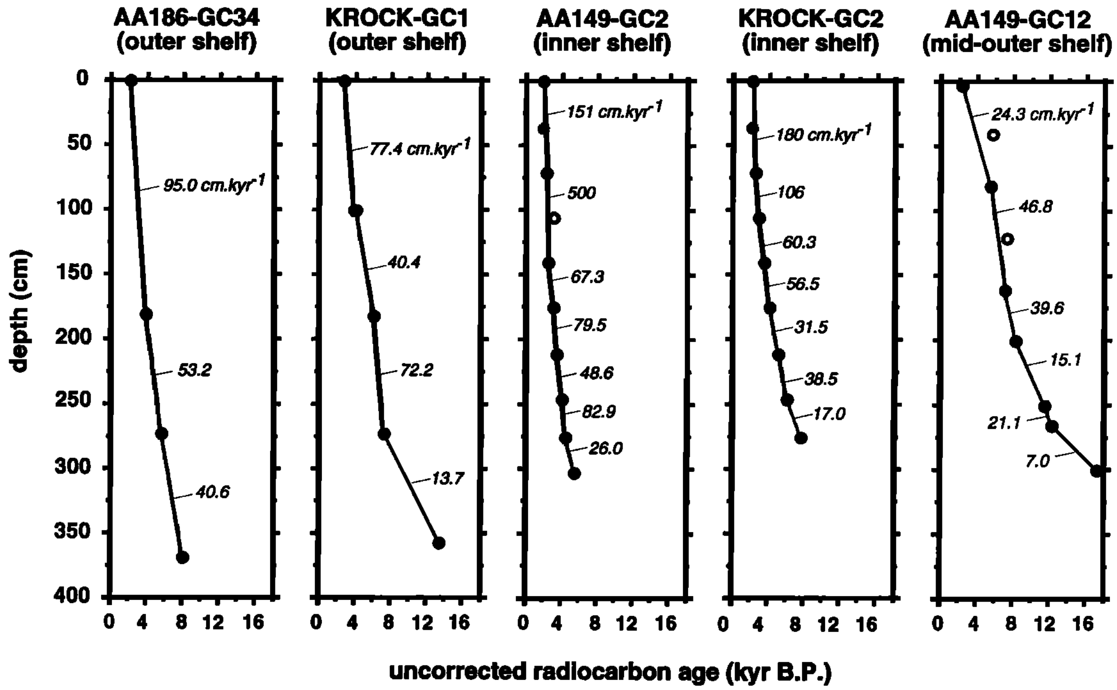


Figure 3. Uncorrected radiocarbon ages versus depth and calculated sediment accumulation rates between dated samples. Samples were excluded from accumulation rate calculations where (1) a sample is older than that immediately below it (open circles) or (2) a sample age is within analytical uncertainty of the core top sample age (stippled circles).

carbon ages of the core top samples range from ~50 to 900 radiocarbon years, which may reflect the previously mentioned uncertainties in our age correction, as well as some loss of core top material, although the latter is only likely to be significant (i.e., greater than ~100-200 years) for cores AA186-GC34, KROCK-GC1, and KROCK-GC2, on the basis of our excess ^{210}Pb analyses. The corrected core top age of KROCK-GC1, if due solely to core top loss, suggests that ~70 cm of the uppermost sediment column is missing, using our estimated accumulation rates. However, as noted above, these accumulation rate estimates and thus estimates of core top losses are very poorly constrained.

Core chronologies were established by linear interpolation of the age versus depth data as shown in Figure 3, using the corrected radiocarbon ages presented in Table 2. With the exception of core AA149-GC12 (see below), the down core physical data (Figure 2) and X-ray images provide no clear evidence of graded beds or erosional surfaces within the SMO units, and coarse material typical of shallower areas on the shelf, including calcareous biota which occur on nearby Fram Bank [*Rathburn et al.*, 1997], is rare. This and the absence of significant age reversals or hiatuses in the down core radiocarbon ages argues against significant episodic down slope transport of sediments into these basins, or significant erosional events, although the sustained transport of fine sediments from the shallow shelf areas into these basins is clearly an important process [*Harris and O'Brien*, 1996, 1998]. Thus, in our interpretation of the down core geochemical data we assume that each core contains a continuous record of sediment accumulation that has not been significantly disturbed by slumps, turbidity flows, or erosional events. In the case of core AA149-GC12, however, X-ray images indicate ripple cross bedding concentrated between 50 and 150 cm depth, which along with radiocarbon age reversals (Figure 3) and relatively depleted $\delta^{13}C$ values ($\leq$ 30‰) between 40 and 120 cm depth (Table 2), suggest that the sediment record in this core may reflect episodes of down slope sediment transport by relatively strong density currents [*Harris*, 2000].

3.2. Physical and Geochemical Data

The SMO units are made up of ~70-95% mud and ~5-30% sand (Figure 2) and contain 16-46% opal (mean is 37%) and 0.66-2.3% TOC (mean is 1.3%) by weight. Down core magnetic susceptibility values are uniformly low (< 5 x 10^{-6} cgs) and featureless within the SMO facies (Figure 2), indicating a paucity of ferromagnetic minerals and lithogenic material in general. Microscopic examination of core material indicates that the majority of organic matter is associated with the remains of diatoms that are typical of Antarctic coastal and shelf waters, with lithogenic particles (fine sand to coarse silt) accounting for the remainder of the SMO facies material. A detailed microfossil study of cores KROCK-GC1 and KROCK-GC2 [*Taylor*, 1999; F. Taylor and A. McMinn, Evidence from diatoms for Holocene climate fluctuation along the East Antarctic margin, submitted to *The Holocene*, 2000] indicates that the diatoms *Fragilariopsis curta* and *Fragilariopsis cylindrus* dominate the dark olive green SMO, except near the base of both cores where *Chaetoceros* spp. resting spores are dominant, whereas the lighter-colored bands in core KROCK-GC1 are characterized by an increased abundance of the diatom *Corethron criophilum*. Preliminary microscopic examination of material from core AA186-GC34 suggests that the light and dark banding in this core reflects diatom species assemblages similar to those in core KROCK-GC1 (L. Armand, personal communication, 1999). The occurrence of the larger visible "*Corethron* layers" in these cores is indicated in Figure 2.

In examining the paleoenvironmental record preserved in these cores, specifically the record of accumulation of biogenic versus lithogenic material, the physical measurements are of only limited use at the level of resolution used in this study because of the relatively homogeneous grain size distribution and the relatively low and constant values of magnetic susceptibility. We have therefore focused on chemical proxies for the accumulation of lithogenic and biogenic material in an effort to infer the sedimentation histories for these core sites. The chemical data considered here are divided into the following groups on the basis of their utility as paleoenvironmental proxies.

1. Ti, Al, and Fe, the major fractions of which are generally associated with lithogenic material in pelagic and hemipelagic sediments [*Calvert and Pedersen*, 1993; *Kumar et al.*, 1995]. Bulk sediment concentrations of these elements thus provide a relative measure of the accumulation of lithogenic material.

2. TOC and Br, which provide relative measures of the accumulation of organic matter in marine sediments. TOC provides a direct measure of the deposited organic matter remaining after diagenetic remineralization, whereas Br is thought to be uniquely controlled by the organic fraction in marine sediments [*Price et al.*, 1970; *Calvert and Pedersen*, 1993].

3. Opal and total Si/Al, both of which provide relative measures of the accumulation of biogenic siliceous matter where postdepositional preservation is high or relatively constant [*Charles et al.*, 1991; *Mortlock et al.*, 1991; *McManus et al.*, 1995]. Opal determined by the method of *Mortlock and Froelich* [1989] provides a direct measure of this biogenic silica, whereas total Si (determined by X-ray fluorescence), when normalized to Al, provides an indirect estimate of this same quantity, assuming that the majority of Al is lithogenic.

4. Mo, U, and Mn, which change valence and are thus adsorbed or precipitated in response to changes in sedimentary redox conditions. Sedimentary enrichments in Mo/Al, U/Al, and Mn/Al, relative to crustal abundances (assuming all Al is lithogenic), are indicative of anoxic (sulfidic), suboxic, and oxic conditions, respectively [*Calvert and Pedersen*, 1993, 1996; *Crusius et al.*, 1996]. Thus these elements serve as sensitive proxies of sedimentary redox conditions, with down core changes in Al-normalized concentrations indicating past variations in redox conditions, as may result from variations in the accumulation of organic matter.

5. Ni, Cu, and Zn, which form insoluble metal sulfides and may thus precipitate in reducing sediments where dissolved sulfide is present [*Calvert and Pedersen*, 1993]. Sedimentary enrichments in Ni/Al, Cu/Al, and Zn/Al relative to crustal abundance (assuming all Al is lithogenic) are indicative of anoxic (sulfidic) conditions, as may result from increased accumulation of organic matter.

6. Biogenic (or excess) Ba, calculated as the difference between total and lithogenic Ba, where lithogenic Ba is estimated from Al using the average crustal weight ratio of Ba/Al (0.0075). Biogenic Ba is thought to be delivered to the sediments as barite contained in organic matter from the surface ocean and is well preserved in oxic sediments, where it serves as a proxy for export production in overlying waters [*Dymond et al.*, 1992].

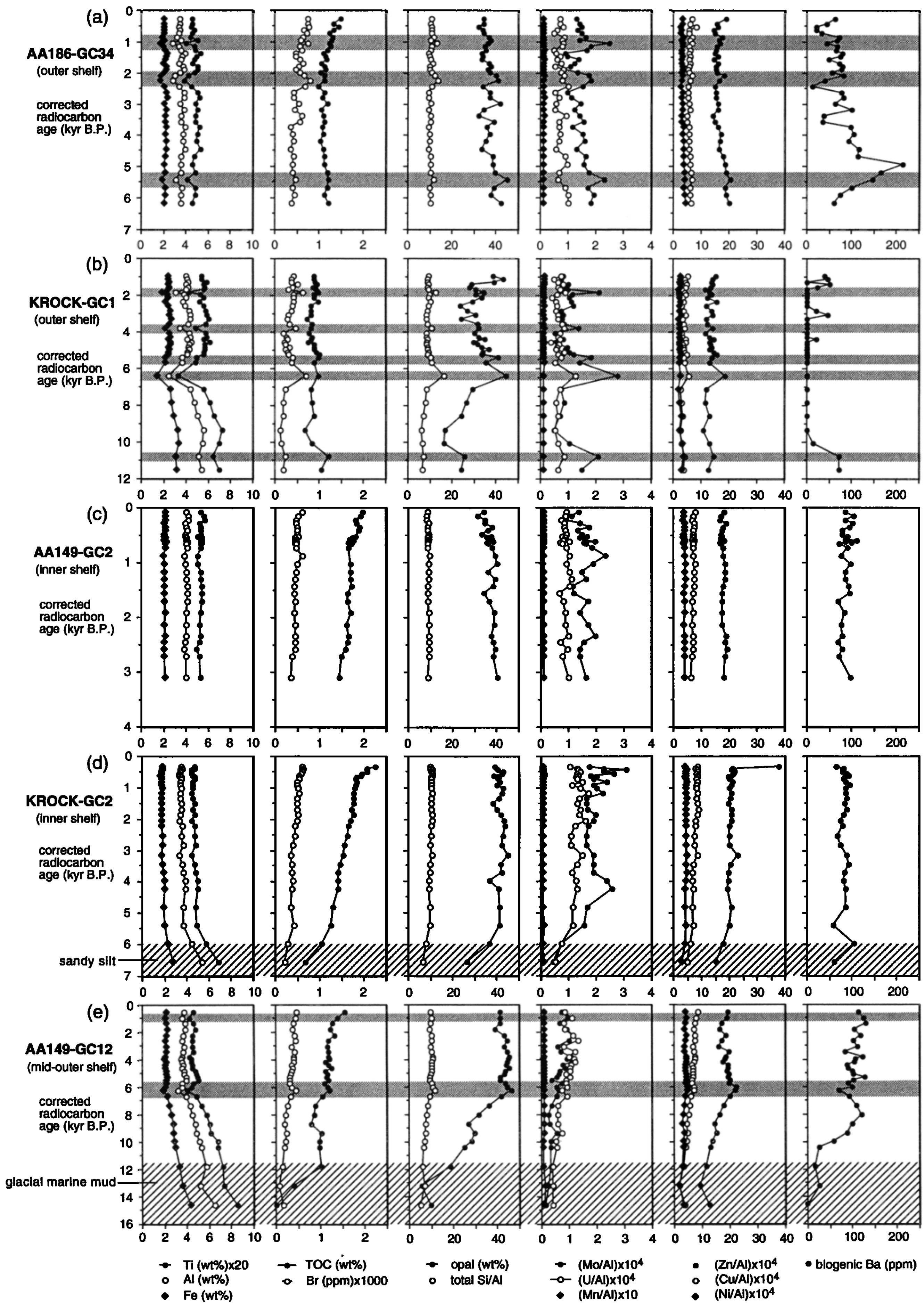
These chemical data are presented in Figure 4, in which bulk sediment concentrations (dry weight basis) are plotted against the age corresponding to the average depth of each subsample in the core. These ages were linearly interpolated using the chronological scheme described in section 3.1. The periods of deposition integrated by these 10 cm thick subsamples range from ~20 years (base of AA149-GC12) to 1400 years (upper portion of AA149-GC2), based on the estimated sediment accumulation rates shown in Figure 3, with a typical temporal resolution of ~200 years for a sediment accumulation rate of 50 cm kyr^{-1}.

4. Discussion

4.1. Down Core Compositional Records

The SMO units of the five cores are generally similar in terms of bulk composition, with cores KROCK-GC1 and AA149-GC12 displaying the greatest compositional ranges, as is evident in the down core opal data (Figure 4). Manganese concentrations are uniformly low in all of the cores, with Mn/Al values (mean is 0.16 ± 0.27) statistically indistinguishable from the average shale ratio of 0.08 [*Wedepohl*, 1971], whereas the ratios Mo/Al (mean is $1.3 \pm 0.6 \times 10^{-4}$) and U/Al (mean is $0.9 \pm 0.3 \times 10^{-4}$) are significantly enriched relative to the average shale ratios of ~0.3×10^{-4} and ~0.4×10^{-4}, respectively [*Wedepohl*, 1971]. These geochemical trends together with shipboard observations of hydrogen sulfide odor upon recovery of these cores and grab sample AA186-GB9 indicate that the entire sediment column and possibly the sediment-water interface were anoxic (sulfidic) at each of these sites. This would act to limit bioturbation by benthic organisms, thus favoring the preservation of high-resolution records of sediment accumulation. However, the presence of hydrogen sulfide in the sediment column also precludes the use of biogenic Ba as a productivity proxy, since dissolution of sedimentary barite accompanies sulfate reduction [*Dymond et al.*, 1992]. Indeed, the down core records show that biogenic Ba does not vary in concert with the other proxies of organic matter accumulation, with some calculated values being close to or less than zero (Figure 4). A similar situation is likely to apply in sediments from other basins on the Antarctic margin, such as the Palmer Deep, where sedimentary Al and Ba concentrations [*Rodriguez and Domack*, 1994] sometimes yield negative values for biogenic Ba.

The geochemical records of the two outer shelf cores AA186-GC34 and KROCK-GC1 (Figures 4a and 4b) show several prominent minima in the lithogenic elements Ti, Al, and Fe, which are generally coincident with small maxima in the biogenic components TOC, Br, opal, and Si/Al and also coincident with maxima in Mo/Al and, in some cases, U/Al and Zn/Al. These features are indicated by the stippled bands in Figure 4, each of which is 500 years in thickness. We suggest 500 years as a conservative estimate of the uncertainty in the timing of these chemical features, given that the subsamples used for geochemical analysis typically integrate several centuries of accumulation and the chronological uncertainties of 60-280 years in our radiocarbon dates. These minima in lithogenic components are also generally coincident with down core minima in sand content (Figure 2) and maxima in water content (data not shown). In core AA186-GC34, all of these features coincide with light-colored *Corethron* layers, whereas for KROCK-GC1, three of five lithogenic minima coincide with *Corethron* layers, while a fourth occurs in the *Chaetoceros* layer near the bottom of the core. These observations immediately suggest that the down core



(a)
AA186-GC34
(outer shelf)
corrected radiocarbon age (kyr B.P.)
(b)
KROCK-GC1
(outer shelf)
corrected radiocarbon age (kyr B.P.)
(c)
AA149-GC2
(inner shelf)
corrected radiocarbon age (kyr B.P.)
(d)
KROCK-GC2
(inner shelf)
corrected radiocarbon age (kyr B.P.)
sandy silt
(e)
AA149-GC12
(mid-outer shelf)
corrected radiocarbon age (kyr B.P.)
glacial marine mud
Ti (wt%)x20
Al (wt%)
Fe (wt%)
TOC (wt%)
Br (ppm)x1000
opal (wt%)
total Si/Al
(Mo/Al)x10^4
(U/Al)x10^4
(Mn/Al)x10
(Zn/Al)x10^4
(Cu/Al)x10^4
(Ni/Al)x10^4
biogenic Ba (ppm)

Figure 4.

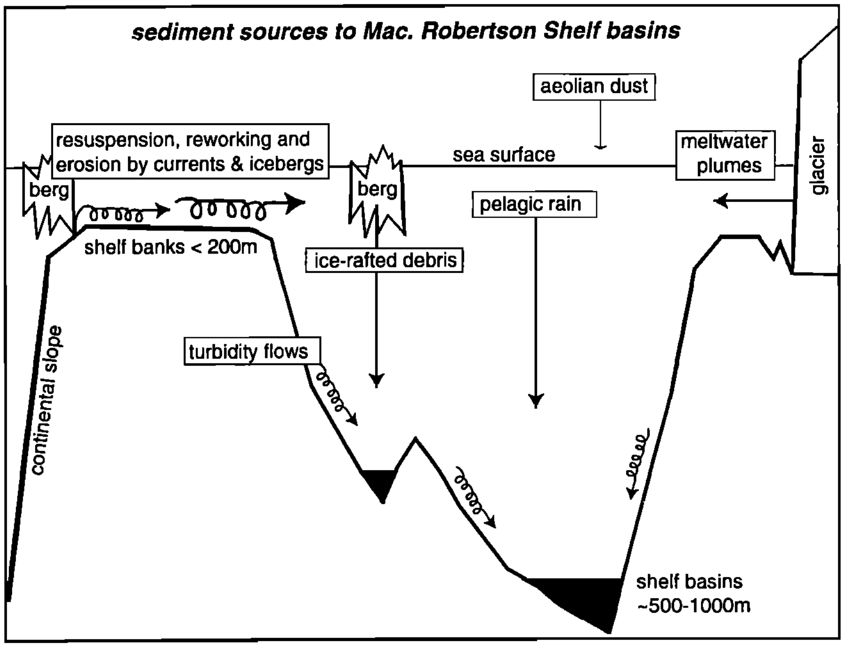


Figure 5. Major sediment sources to the Mac. Robertson Shelf basins. Modified from Figure 10 of *Harris and O'Brien* [1996], © Springer-Verlag 1996.

minima in lithogenic elements may record enhanced deposition of pelagic biogenic material on the surrounding shelf, perhaps due to blooms of *Corethron* or *Chaetoceros* in these waters.

The inner shelf cores AA149-GC2 and KROCK-GC2 contain generally finer-grained sediments (Figure 2) and higher concentrations of TOC, opal, and Mo (Figures 4c and 4d) than the outer shelf cores, suggesting either enhanced deposition of biogenic material relative to lithogenic material in the inner shelf basins or enhanced preservation of organic matter in these locations, which would create more reducing conditions in the sediment column. In contrast to the outer shelf cores, the down core records of the two cores from the inner Nielsen Basin contain no significant minima in lithogenic components. Rather, the relative proportions of lithogenic and biogenic material which has accumulated at these sites has been remarkably uniform during the middle to late Holocene, before which this area was probably covered by a floating ice shelf [*Harris and O'Brien*, 1998]. In both cores, there are significant decreases in TOC content with increasing sediment age, which may reflect an increase in the accumulation of organic carbon during the late Holocene or, more likely, since there are no corresponding trends in biogenic silica, the progressive diagenetic remineralization of TOC within the sediment column [*Berner*, 1980; *Domack and McClennen*, 1996].

Not surprisingly, the chemical record of core AA149-GC12 (Figure 4e), from midshelf of the Nielsen Basin, is generally intermediate between the inner and outer shelf core records. Minima in lithogenic components are discernible at ~6 kyr B.P. and possibly ~1 kyr B.P., as are small maxima in the biogenic components, and there is a steady decrease in the proportion of lithogenic components from ~11 to 6 kyr B.P. In contrast to the other cores, Mo/Al values are less than U/Al, suggesting lower concentrations of hydrogen sulfide, thus less reducing conditions, within the sediment column at this site. This observation suggests a slower accumulation of organic matter at this location during the Holocene compared with the other core sites and is consistent with the sediment accumulation rates calculated for AA149-GC12, which are generally less than those calculated for contemporaneous sections of the other cores in this study (Figure 3). This likely reflects a lesser degree of sediment focusing (see below) at the site of core AA149-GC12, which is not within the central axis of the Nielsen Basin (Figure 1), as well as episodic ventilation by sinking, oxygenated surface waters in this area of the shelf [*Harris*, 2000].

4.2. Paleoenvironmental Interpretation

In order to interpret our proxy records for the accumulation of biogenic and lithogenic material we must first consider the sources of the SMO facies sediments in the Mac. Robertson Shelf basins. The major sources of sediments in these basins are shown schematically in Figure 5, as proposed by *Harris and O'Brien* [1996]. Under seasonally open marine conditions the basins will have received sediments from overlying waters in the form of pelagic rain (biogenic material), ice-rafted debris (lithogenic material), glacial meltwater plumes (lithogenic material), and aeolian dust (lithogenic material). In addition, a mixture of biogenic and lithogenic material will have been transported laterally into the basins from the shallower areas of the shelf as a result of

Figure 4. Down core geochemical data for the five cores in this study. Stippled bands indicate inferred production episodes; hatching indicates sandy silt or glacial marine facies. Estimated analytical uncertainties (± 2σ) shown by width of symbols/bars along bottom axis.

resuspension and erosion by currents, icebergs, and turbidity flows. *Harris and O'Brien* [1996] have noted that net erosional conditions exist over ~90% of the Mac. Robertson Shelf, where coarse-grained sediments dominate the shallow banks and slopes, whereas net depositional conditions exist on the remaining 10% of the shelf area, as represented by the accumulation of fine-grained SMO deposits in the shelf basins. A similar sedimentation pattern has been described for other parts of the Antarctic margin, including the Ross Sea, the Northern Victoria Land Shelf, and the Wilkes Land Shelf [*Anderson et al.*, 1984; *Dunbar et al.*, 1985; *Anderson*, 1999].

The general absence of coarse lithogenic material, graded beds, and erosional surfaces in our cores argues against significant lateral transport of sediments into the shelf basins by icebergs, slumps, and turbidity flows, except in the case of core AA149-GC12, which shows evidence of episodic down slope sediment transport by density currents (as described in section 3.1 and *Harris* [2000]). In addition, swell waves and tidal currents are thought to play a minor role in reworking shelf sediments in our study region [*Harris and O'Brien*, 1998]. On the basis of these considerations and the observations of relatively strong bottom currents on the Mac. Robertson Shelf [*Harris and O'Brien*, 1998] we suggest that the shelf-basin SMO deposits are dominated by fine-grained material which has been winnowed from the shallow banks and slopes, mainly by the action of currents, with a lesser contribution of sediments derived from overlying waters. The major current involved in this process would be the Antarctic Coastal Current, a strong, circumpolar current driven by the East Wind Drift and the density gradient across the Antarctic Slope Front [*Smith et al.*, 1984; *Harris and O'Brien*, 1998; *Bindoff et al.*, 2000]. Current measurements from the Mac. Robertson Shelf suggest that this current flows consistently toward the west, with maximum speeds of 50-200 cm s^{-1} on the outer shelf, with some reduction in strength during the winter and spring [*Harris and O'Brien*, 1998]. Such current speeds would certainly be sufficient to mobilize mud and fine sand from the shallow banks and slopes, which would be re-deposited in the downstream shelf basins and would account for the accumulation of finer sediments in basins on the inner shelf, where current speeds are less.

This west flowing current regime has probably existed on the Mac. Robertson Shelf during most of the Holocene, so that there has been a roughly continuous transport of fine biogenic and lithogenic material (i.e., SMO) into these basins from the shallow shelf areas to the east. Thus the undisturbed sediments which have accumulated in these basins during the Holocene will most likely provide us with a record of the production and deposition of fine sediments, both biogenic and lithogenic, over larger areas of the adjacent, upstream banks. This scenario provides a likely explanation for the apparent increases in sediment accumulation rates during the course of the Holocene that are indicated by the data in Figure 3. *Harris and O'Brien* [1998] have argued that the glacial ice sheet began to retreat from its grounding position on the middle to outer shelf at ~10-12 kyr B.P., and that the calving front of the floating ice sheet had retreated to the inner shelf, near the location of KROCK-GC2, by ~6 kyr B.P. If so, then the seasonally ice-free area of the Mac. Robertson Shelf would have increased significantly between the early and late Holocene, which would be expected to allow the current-driven transport of fine material into the shelf basins from progressively larger "catchment areas" on the shelf. In the remaining discussion, we assume that the SMO sequences contained in our five sediment cores provide a record of the deposition and erosion of fine sediments on the shallow shelf areas immediately to the east, with the exception of the section of core AA149-GC12 that is thought to be disturbed by down slope sediment transport.

What processes might explain the down core variations in the proportions of lithogenic and biogenic components preserved in our middle and outer shelf cores? There are two likely alternatives. The first is that these variations may reflect changes in the current-driven transport of biogenic and lithogenic material into the basins from the shallow areas of the shelf, such that the lithogenic minima in our cores may record periods of lower current speeds, which favor the transport of only the finer biogenic material, such as diatom frustules, from the shelf banks into the basins. This would lead us to expect significantly lower sediment accumulation rates in association with these lithogenic minima; however, we see no evidence for such variations in accumulation rates. Moreover, it is difficult to explain the corresponding down core variations in diatom species abundances, specifically the occurrence of *Corethron*-rich layers, as the result of changes in current-driven resuspension. A more likely alternative is that the down core variations in the proportions of lithogenic and biogenic material may reflect changes in the production and deposition of biogenic material over the shallow shelf areas, such that the lithogenic minima may record periods of enhanced diatom production in shelf waters upstream of the core sites.

A similar interpretation of down core minima in lithogenic components and corresponding variations in diatom species assemblages has been adopted in other studies of sediment cores from Antarctic shelf basins [e.g., *Domack et al.*, 1993; *Leventer et al.*, 1996; *Leventer and Dunbar*, 1996; *Domack and Mayewski*, 1999; *Domack et al.*, 2001]. Thus we interpret the down core minima in lithogenic components in our middle and outer shelf cores as indicative of "production episodes," representing sustained periods of high export production by diatoms in the shelf waters immediately to the east of these shelf basins. The increased accumulation of organic matter in the shelf basins associated with such production episodes would be expected to create more reducing, sulfide-rich conditions within the sediment column, as suggested by the Mo/Al, U/Al and Zn/Al maxima, which approximately coincide with minima in lithogenic components. That most of these chemical features roughly coincide with *Corethron* layers in cores AA186-GC34 and KROCK-GC1 suggests that these production episodes entailed massive blooms of this diatom species in the outer shelf region, given that *Corethron criophilum* is a relatively lightly silicified diatom species that is thought to be poorly preserved in seafloor sediments [*Jordan et al.*, 1991; *Leventer et al.*, 1993, 1996].

4.3. Timing and Forcing of Production Episodes

We assume that the transport of fine sediments from the shallow shelf areas into the basins is relatively rapid, so that within the resolution of our radiocarbon chronologies the compositional changes preserved in the middle and outer shelf cores are coeval with depositional changes over the upstream areas of the Mac. Robertson Shelf. However, the actual duration of the production episodes inferred from these core records is

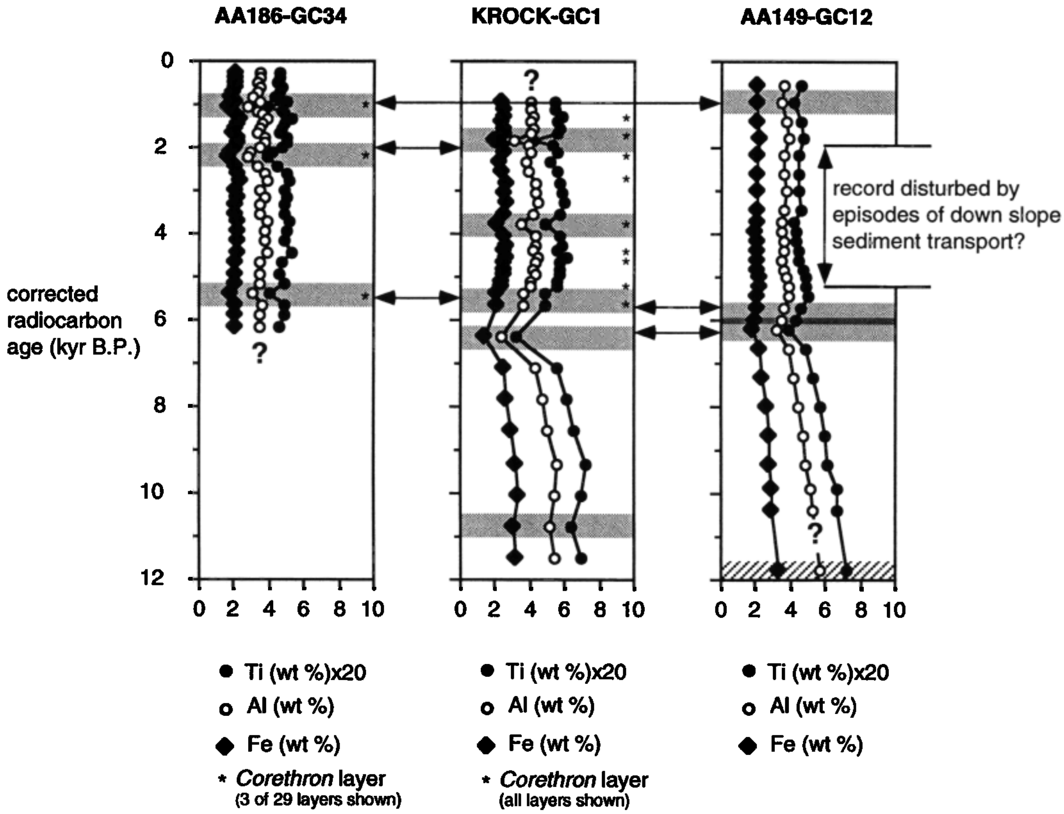


Figure 6. Comparison of timing of production episodes (stippled bands) recorded in the outer and middle shelf cores.

highly uncertain. On the basis of our rather coarse-resolution radiocarbon chronologies, these production episodes lasted of the order of 100-1000 years, although such sediment layers might have accumulated over much shorter periods, given their relatively high water content and our limited stratigraphic resolution. We also note that many of the light-colored layers in the outer shelf cores are only ~1-3 cm in thickness, in which case it would be unlikely that such features would be resolved by our geochemical data, which apply to 10 cm thick down core subsamples. Thus the production episodes which we have inferred from our down core composition data probably represent only the more sustained and/or frequent episodes of export production in adjacent shelf waters.

In Figure 6, we compare the approximate timing of these production episodes in cores AA186-GC34, KROCK-GC1, and AA149-GC12, as indicated by concentration minima in Ti, Al, and Fe and the occurrence of *Corethron* layers in the Iceberg Alley cores. Relative to our corrected radiocarbon chronology, production episodes are indicated at around 1, 2, and 6.2 kyr B.P. in two of the three records, whereas all three core records indicate an episode at ~5.5 kyr B.P. Only core KROCK-GC1 shows evidence of production episodes at ~3.8 and 10.8 kyr B.P., although these episodes may be obscured or unresolved in core AA149-GC12 and the AA186-GC34 record extends back only as far as ~6.2 kyr B.P. The only other reported Holocene sediment records from this region of the East Antarctic shelf are those inferred from microfossil assemblages in two cores from Fram Bank, ~100 km to the east of the Nielsen Basin, where there is evidence of enhanced export production during the period ~2.6-3.4 kyr B.P. [*Rathburn et al.*, 1997]. This period of higher production does not correspond with the timing of the production episodes in our Mac. Robertson Shelf records; however, we note that *Rathburn et al.* [1997] corrected their radiocarbon ages by subtracting 1300 radiocarbon years, accounting only for the reservoir effect. Applying a larger radiocarbon age correction of the order of 1700 years (see section 3.1) to the raw radiocarbon dates of *Rathburn et al.* [1997] places the Fram Bank high-production period at ~2.2-3 kyr B.P., which then overlaps with the ~1.6-2.5 kyr B.P. production episodes indicated in our Iceberg Alley cores (Figure 6).

On the basis of comparisons of sediment records from the Antarctic Peninsula shelf with Northern Hemisphere paleoclimate records it has been suggested that periods of elevated productivity inferred from the Antarctic marine records reflect global warm cycles with periods of ~400, 200, and 50-70 years superimposed upon longer-term periods of low-productivity corresponding to global cooling events [*Leventer et al.*, 1996; *Domack and Mayewski*, 1999; *Domack et al.*, 2001]. To date, the most detailed Holocene sediment records from the Antarctic shelf region are those preserved in cores collected from the Palmer Deep, off the Antarctic Peninsula, during Ocean Drilling Program Leg 178. High-resolution paleoproductivity records have been derived from these cores using down core magnetic susceptibility measurements and radiocarbon dating [*Domack et al.*, 2001]. The production episodes inferred from our Mac. Robertson Shelf cores AA186-GC34, KROCK-GC1, and AA149-GC12 (Figure 6) are roughly coeval with high-productivity periods inferred from the Palmer Deep records [*Domack et al.*, 2001] as well as a mid-Holocene productivity high inferred from Ross Sea sediment records [*Jacobson*, 1997; *Frignani et al.*, 1998; *Cunningham et al.*, 1999], suggesting that there may have been circum-Antarctic periods of enhanced export production during the Holocene in response to global climatic forcing. However, such comparisons must be made with caution, and the apparent agreement between core records from the Mac. Robertson Shelf, the Palmer Deep, and the Ross Sea may be simply fortuitous, given the uncertainties in our radiocarbon chronology and the limited temporal resolution of our geochemical data.

What were the immediate forcing mechanisms responsible for these inferred episodes of enhanced production on the Mac. Robertson Shelf, and why are no such production episodes indicated by the inner shelf cores? The factors which control algal export production and community structure in Antarctic waters are not well understood but are likely to include vertical stability of the upper water column [*Smith and Nelson*, 1990; *Sakshaug et al.*, 1991; *Arrigo et al.*, 1998a, 1999], seeding by sea ice algae [*Smith and Nelson*, 1986; *Leventer and Dunbar*, 1996], grazing by zooplankton [*Lancelot et al.*, 1993; *DiTullio and Smith*, 1996], light limitation due to sea ice cover and self shading [*Smith et al.*, 1996], availability of micronutrient elements such as iron [*Martin et al.*, 1990; *Sedwick and DiTullio*, 1997; *Sedwick et al.*, 2000], and more rarely, depletion of macronutrients [*Tréguer and Jacques*, 1992]. In studies of sediment records from the Antarctic Peninsula continental shelf, periods of enhanced organic matter accumulation have been attributed to enhanced biological production during periods of warm and/or less windy climate, as a consequence of decreased sea ice cover and increased water column stratification, the latter resulting from decreased wind mixing, warmer sea surface temperatures, and increased meltwater inputs [e.g., *Leventer et al.*, 1996; *Domack and Mayewski*, 1999]. In addition, an increased abundance of *Corethron criophilum* and *Chaetoceros* resting spores in sediments from the Ross Sea and the Antarctic Peninsula shelf has been ascribed to enhanced water column stratification [*Leventer et al.*, 1993, 1996].

In considering possible causes for the productivity episodes recorded in our cores from the Mac. Robertson Shelf, we note that most of these features may reflect massive *Corethron* blooms in adjacent shelf waters, with the exception of the production episode recorded at ~10.8 kyr B.P. in core KROCK-GC1, which apparently records production of *Chaetoceros* resting spores associated with the retreat of a permanent ice cover. In contrast, the sediment cores from the inner Nielsen Basin suggest relatively uniform export production in upstream shelf waters. With the exception of the base of core KROCK-GC2, biogenic material in these SMO sequences is dominated by the diatom *Fragilariopsis curta*, which is thought to be favored in ice-marginal environments [*Leventer et al.*, 1993]. Satellite observations obtained in recent years have revealed the existence a persistent coastal polynya on the inner Mac. Robertson Shelf, the Cape Darnley polynya, which is probably maintained by the presence of icebergs grounded on the shallow banks off Cape Darnley (~70°E) [*Massom et al.*, 1998]. Satellite advanced very high resolution radiometer images obtained during 1997-1999 (see http://www.antcrc.utas.edu.au/avhrr/mawson/archive/) reveal that the ice-free waters of this polynya extend west from Cape Darnley as far as the inner Nielsen Basin during the spring and summer, whereas the outer shelf waters are covered by pack ice during the spring and sometimes into the summer. If we assume that a similar pattern of seasonal ice cover has existed since the middle Holocene, by which time the glacial ice shelf had retreated to the inner continental shelf, then the location of the Cape Darnley polynya provides a likely explanation for the differences between the inner and outer shelf core records.

During the middle to late Holocene, the ice-marginal, well-mixed open waters of the Cape Darnley polynya may have favored consistent interannual production by *F. curta* during the spring and summer, which would have resulted in the consistent transport of fine, biogenic-rich sediments into the inner Nielsen Basin by the relatively weak inner shelf bottom currents. In contrast, the often ice-covered waters of the outer Mac. Robertson Shelf may have experienced much more variable algal production during the spring and summer as a result of interannual variations in sea ice cover. Moreover, unlike the wind-mixed open waters of the polynya, the outer shelf waters would be more likely to maintain a stable upper water column as a result of enhanced sea ice melting during warmer years. In this scenario, algal production and community composition upstream of the outer shelf basins would have been strongly modulated by the extent of sea ice melting during the spring and summer, whereas water column conditions and algal production to the east of the inner Nielsen Basin were "buffered" by the presence of the Cape Darnley polynya. From these considerations we suggest that the production episodes recorded in our outer shelf cores from Iceberg Alley and more weakly in the midshelf core from the Nielsen Basin record years of enhanced sea ice melting during climatic warm periods, when factors such as a meltwater-stratified water column [*Leventer et al.*, 1996] and ice-derived iron inputs [*Sedwick et al.*, 2000] stimulated massive blooms of diatoms, particularly *Corethron criophilum*, on the outer Mac. Robertson Shelf.

4.4. Conclusions and Future Work

Our interpretations of sediment records from the Mac. Robertson Shelf basins are generally consistent with other Holocene sediment records from the Antarctic margin in that they provide evidence of millennial-scale episodes of enhanced diatom production in outer shelf waters, including a mid-Holocene "warm period" between ~5 and 7 kyr B.P. Such production episodes may be related to increased sea ice melting during global climatic warm periods, although local forcing mechanisms cannot be discounted at this time, given the dearth of sediment records from our study area. Our results also indicate that small-scale regional phenomena, such as coastal polynyas, may strongly influence the sediment records in Antarctic shelf basins, which highlights the need for careful consideration of regional sediment dynamics in paleoenvironmental studies of the Antarctic margin. Further studies are clearly required in order to establish the extent and timing of Holocene paleoenvironmental changes on the East Antarctic margin and, at a more basic level, to clarify the relationship between the sediments accumulating in shelf basins and the processes occurring in overlying waters. Such information can only be provided by increasing the spatial coverage and temporal resolution of sedimentary records from this region for comparison with emerging high-resolution paleoenvironmental records from other areas.

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